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UPPER PLEISTOCENE UPLIFTED SHORELINES AS TRACERS OF (LOCAL RATHER THAN GLOBAL) SUBDUCTION DYNAMICS

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ABSTRACT

Past studies have shown that high coastal uplift rates are restricted to active areas, especially in a subduction context. The origin of coastal uplift in subduction zones, however, has not yet been globally investigated. Quaternary shorelines correlated to the last interglacial maximum (MIS 5e) were defined as a global tectonic benchmark (Pedoja et al. (2011)). In order to investigate the relationships between the vertical motion and the subduction dynamic parameters, we cross-linked this coastal uplift database with the "geodynamical" databases from Heuret (2005), Conrad and Husson (2009) and Müller et al. (2008). Our statistical study shows that: [1] the most intuitive parameters one can think responsible for coastal uplift (e.g., subduction obliquity, trench motion, oceanic crust age, interplate friction and force, convergence variation, dynamic topography, overriding and subducted plate velocity) are not related with the uplift (and its magnitude); [2] the only intuitive parameter is the distance to the trench which shows in specific areas a decrease from the trench up to a distance of ~300 km; [3] the slab dip (especially the deep slab dip), the position along the trench and the overriding plate tectonic regime are correlated with the coastal uplift, probably reflecting transient changes in subduction parameters. Finally we conclude that the first order parameter explaining coastal uplift is small-scale heterogeneities of the subducting plate, as for instance subducting aseismic ridges. The influence of large-scale geodynamic setting of subduction zones is secondary.

Highlights

- Large-scale geodynamics only explain first order coastal uplift rates
- Uplift is localized over asperities of the subducting plate
- Uplift could be related to transient changes in subduction parameters

- Rapid uplift is restricted to the area <300 km from the trench (forearc)
- Forearc and plate interiors respond differently

Keywords: marine strandline; shoreline; uplift; subduction; geodynamics; Quaternary

1. INTRODUCTION

Fossil shorelines (or strandlines) are generally packed and constitute staircase coastal geomorphologies or sequences of "terraces" (e.g. marine or reefal for example). They are tracers of the sea level at the time they formed. Current elevation of fossil shorelines results from the combination of sea level change (eustasy) and vertical ground motion (uplift or subsidence, Lajoie et al. (1991), Pirazzoli et al. (1993)). Pedoja et al. (2011) exhaustively compiled the worldwide repartition and elevation of the shorelines formed during the last interglacial sea level highstand (Marine Isotopic Stage 5e, ~120 ka BP) and calculated apparent coastal uplift rates since that time. More recently, Pedoja et al. (in press), investigated other benchmarks (MIS 1, 3, 11 and upper shoreline of the sequences) in the coastal sequences including MIS 5e strandline. Their database highlights the contrast in tectonic uplift rates between active zones (mainly Pacific Ocean) and passive zones (Atlantic and Indian Oceans) (Figure1). Even if Pedoja et al. (in press) did a first-order exploration of uplift record on paleoshorelines in function of the rough geodynamic setting, vertical motion along the coasts located above subduction zones has never been extensively explored. In this paper, we look for possible geological parameters that may explain why coastal areas located above subduction zones are uplifting so fast (Figure 1).

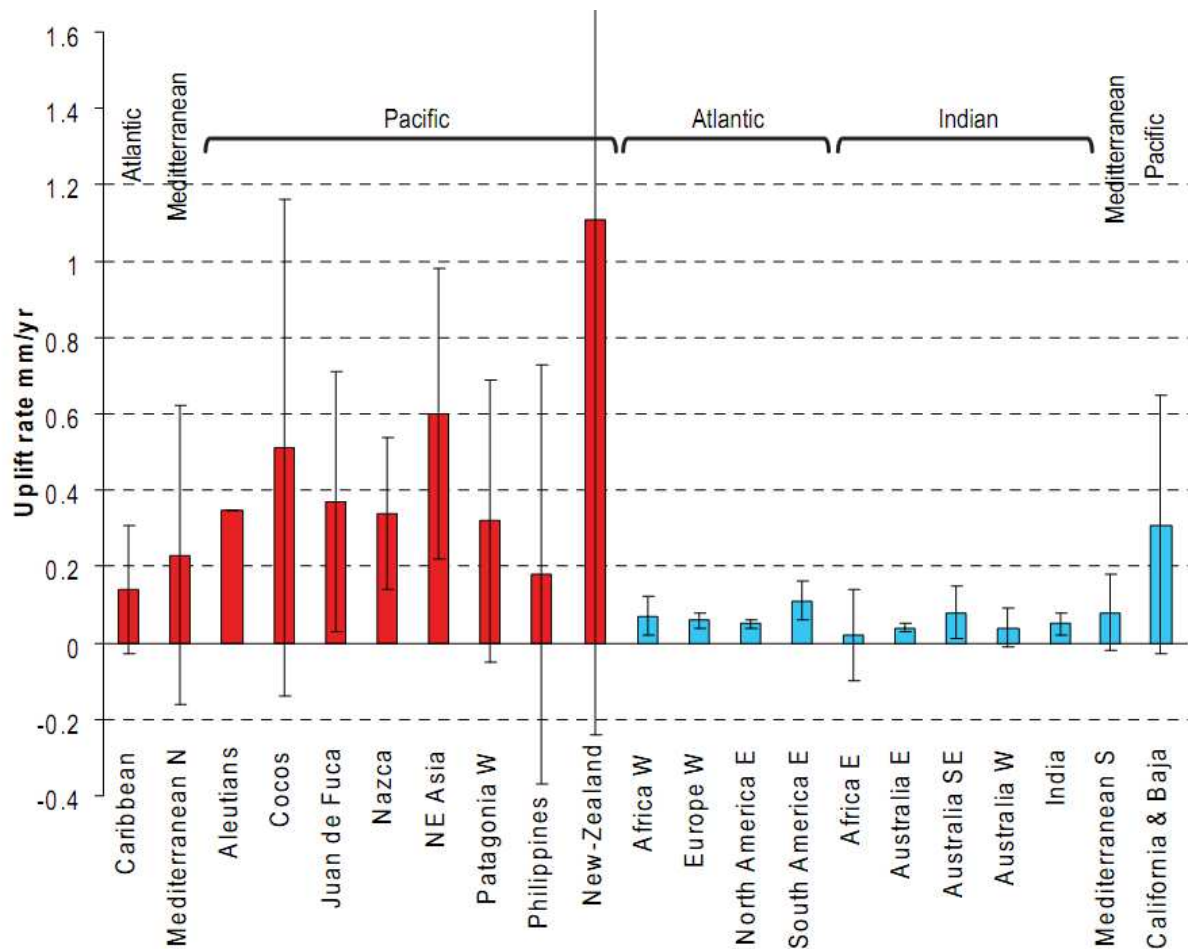


Figure 1. Worldwide distribution of apparent coastal uplift rates (since MIS 5e): in red and blue the average uplift rate for, respectively, the actively deforming zones (mostly subduction zones) and the stable zones (mostly passive margins; data from Pedoja et al. (2011)). Brackets represent the data standard deviation. Note the zone named California and Baja, corresponds to a passive margin very close to a rift/transform setting.

The compilation from Pedoja et al. (2011) records only emerged terraces with few exceptions. As discussed in Pedoja et al. (2011), the worldwide distribution of shoreline sequences suggests that there are much less subsiding areas along subduction coastlines than uplifting ones, a fact that shall not be considered as an observational bias (see Pedoja et al., 2011, 2014). Then, the database may reflect a global tendency to coastal uplift during late Pleistocene (Pedoja et al. (2011)), and also partly results from the fact that Pleistocene to present-day coastal subsidence is more difficult to quantify than coastal uplift. In any case, this database shows that the average coastal uplift is faster above subduction zones than at passive margins. In the following, we look for possible links between Late Pleistocene (posterior to MIS5) coastal uplift and subduction geodynamics. In particular, we investigate the uplift dependence on some geodynamic parameters, chosen for their driving effect. Some are obvious, like: distance to the trench, trench motion, age of the subducting plate, subduction obliquity, overriding and subducting plate velocities, and dynamic topography. The others are suspected to act on the vertical motion but with magnitudes and direction that deserve exploration: interplate force and friction force, position along the trench (i.e. distance to the subducting plate edge), slab dip, tectonic setting of the overriding plate (see Heuret, 2005).

2.1 Databases: paleoshorelines and geodynamics

The compilation by Pedoja et al. (in press); Pedoja et al. (2011) focuses on coastal geomorphic indicators correlated to the Marine Isotopic Stage 5e (125 ky BP). Indeed, corresponding terraces are the most extensively preserved and dated. Moreover, MIS 5e is purportedly the last analogue to the current interglacial and the time span is enough to largely exceed several seismic cycles such that the uplift rate is not significantly affected by an individual seismic event. Using the MIS 5e shoreline elevation, we calculated the average uplift rate using the following formula: $U=(z-e)/t$, with U the shoreline uplift rate, z the MIS 5e terrace elevation, t the age of the terrace and e the relative elevation of the MIS 5e sea level with respect to the current sea level. In accordance to Pedoja et al. (in press); Pedoja et al. (2011)), we use $e= 0\pm 10$ m, which is conservative in the sense it takes into account the different debated evaluations of the last interglacial sea-level (e.g., Waelbroeck et al. (2002), Kopp et al. (2009); O'Leary et al. (2013)) and the way the shorelines are fossilized (e.g., Lajoie et al. (1991)). In addition, this elevation value is of little interest to the current study as it uniformly offsets uplift rates while our analysis considers relative vertical displacements from one site to another. Besides the elevation of the uplifted shorelines, Pedoja et al. (in press; 2011) deliver some additional information like the geographic location of the sequences. Noteworthy, the spatial repartition of the data over South America, Japan and Cascadia subduction allow investigating the coastal uplift distribution as a function of the distance to the trench up to 800 km away (in the Japan and South America transects). In addition, it is noticeable that some places have not been investigated for marine terraces, like the Aleutian subduction zone where the Ostrov Beringa and Segum islands exhibit marine terraces visible on satellite images but not studied in the field or even the Mariana subduction zone (Stafford et al. (2005) observed uplifted karst in Guam).

Subduction zone geodynamic parameters are sourced from Heuret (2005)(parts of the data base have been published in Heuret & Lallemand, 2005, Lallemand et al., 2005 and Funiciello et al. 2008). He provides every 2 degrees multiple geodynamic parameters like the overriding plates tectonic regime, the trench motion, the proximity of the measurement to a subduction edge, the shallow and deep slab dip, the age of the adjacent oceanic crust and the plate convergence rate. In addition, we extracted the current dynamic topography rate of change for the last Myr from the dynamic topography data released by Conrad and Husson (2009) and Müller et al. (2008). Besides end-member models HS3 (Gripp and Gordon 2002) or NNR, most of the reference frames fall in the same range (see e.g. fig.6 in Becker, 2006). Given that, we chose the moving hotspot model of Steinberger et al. (2004) to calculate the trench, the overriding and subducting absolute plate motion because we see it as representative of most frameworks (see Funiciello et al for some comparisons between frameworks). Thus, we emphasize that our observations and conclusions would not be altered by choosing an alternative reference frame. The reference frame defined by Steinberger et al. (2004) is very similar to that of O'Neill et al. (2005) which is based on the Indo-Atlantic hotspots. Such reference frames are most in line with predictions from geodynamic models where subducting plates move trenchward and trenches predominantly retreat and where global mantle viscous dissipation is minimized, while this is not the case for HS3 (Schellart et al., 2008). Second, Indo-Atlantic hotspots reference frames are in better agreement with observed global mantle anisotropy than HS3 (Becker et al., 2008, Kreemer, 2009) and with subducted slab structure than HS3 (Schellart, GRL 2011). The convergence rates are extracted from Heuret (2005) (published in Lallemand et al., 2005).

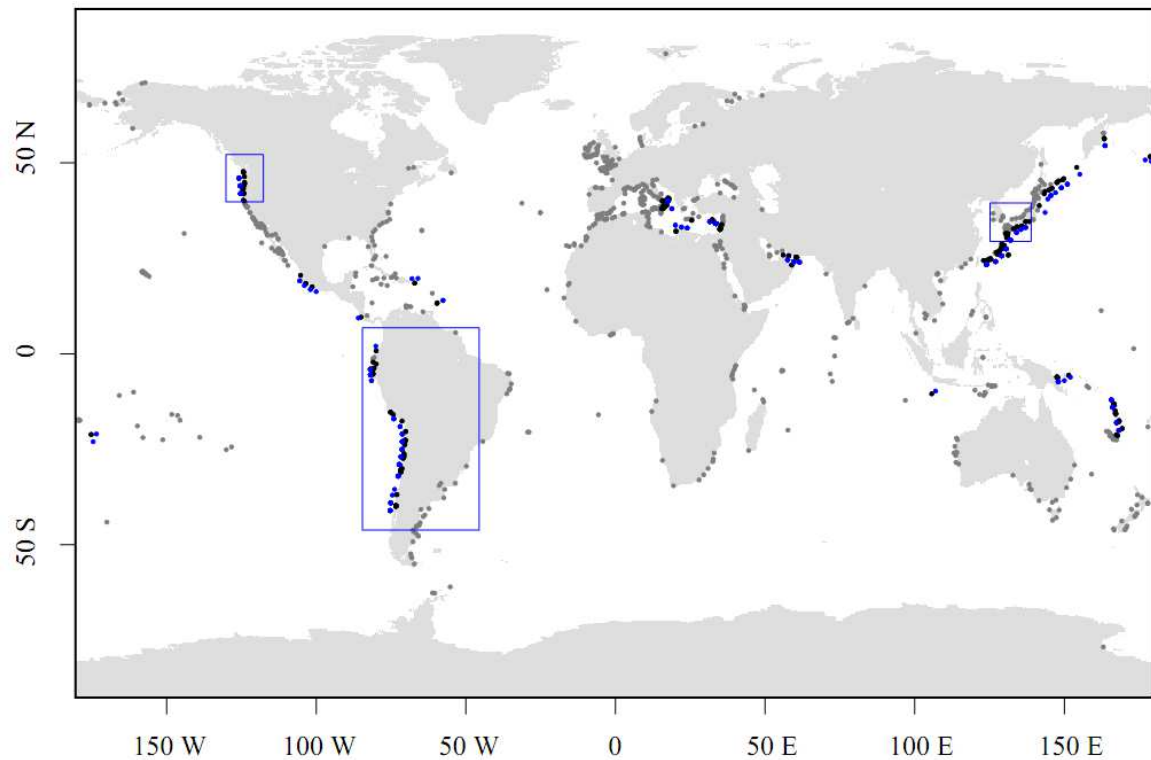
The interplate friction force is calculated after Lallemand (1999) assuming a friction coefficient of 0.3. The interplate force at trench is probably best expressed as the mean trench normal integrated mantle drag called M_d and calculated by Husson (2012). This metric averages the drag forces exerted by the convecting mantle underneath the converging plates. This value is the integral of the normal to trench components of the shear tractions derived from Conrad and Behn (2010) or Conrad and Husson (2009). As, such it is a measure of the net horizontal force that drags plates against each other. Husson (2012) proposed this value to be best correlated to the upper plate tectonic setting (see thereafter, section 3.3).

The Mediterranean Sea has been avoided by Heuret (2005) because subduction zones are small and possibly interact with each other and surrounding collision zones. The dataset of Heuret (2005) is here extended to encompass the Mediterranean Sea (Gibraltar, Tyrrhenian and Aegean subduction zones) and the Makran (Figure 2A).

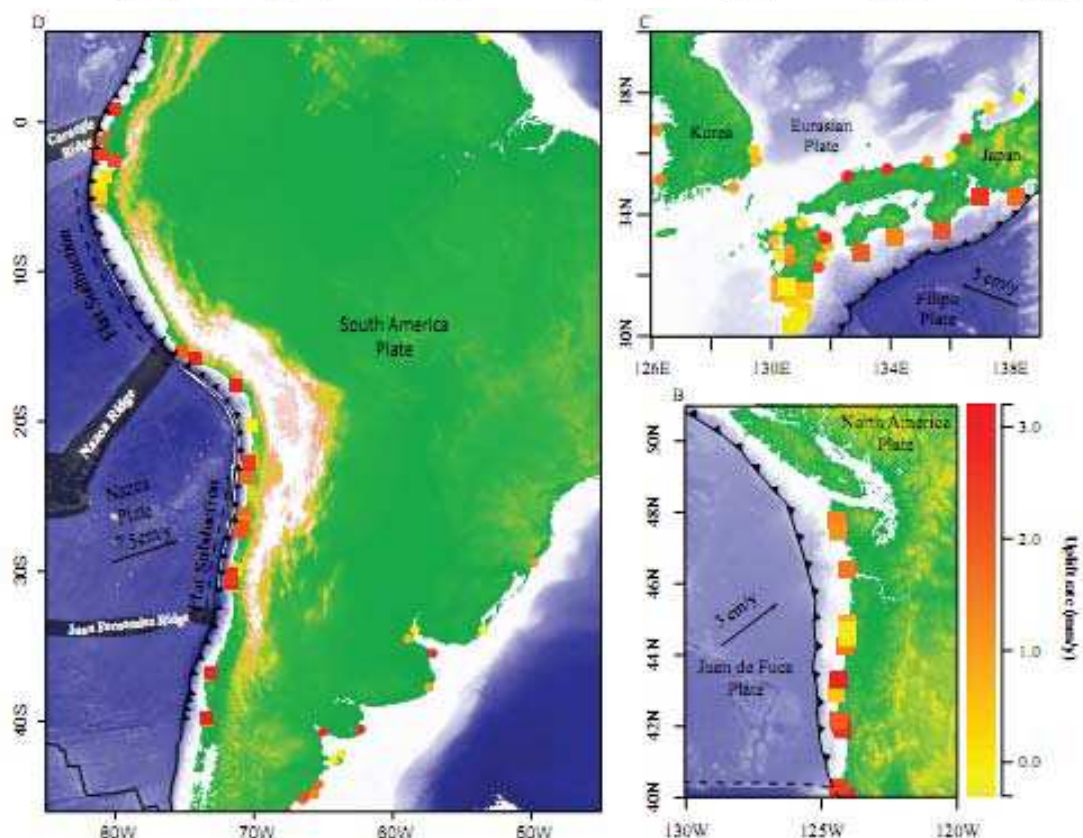
Finally, we combine the 2 datasets (uplifted shorelines from the last interglacial maximum vs. subduction geodynamics), by associating each site of uplifted shoreline to the closest subduction data point in 2 degree-bins. Figure 2 shows the spatial repartition of the uplifted paleocoasts (shoreline, strandline, etc.) and related subduction record. The average uplift rate derived from the marine terraces in this record is ~ 0.2227 mm/y.

In order to define a robust indicator of correlation between the apparent coastal uplift rate and the various geodynamic parameters that we tested, we calculated correlation coefficients r as defined by Pearson (1896):

where x_i and y_i are data coordinates and \bar{x} and \bar{y} their average values. A correlation coefficient of 1 or -1 would respectively indicate a perfect linear correlation or anti-correlation between the uplift rate and the tested parameter. Every data manipulation and calculation has been performed with the software RGui (R Development Core Team (2010)). Note that this correlation coefficient is ineffective to account for non-linear relationships. We will qualify of ‘significant’ every correlation coefficient associated with a p -value < 0.01 . This always comes with a 1σ confidence interval encompassing the 0 value.



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Figure 2. A. Spatial repartition of the two main databases: the MIS 5e uplifted shoreline data from Pedoja et al. (2011) are in grey or black if less than 2 degrees away from any subduction data (Heuret and Lallemand, 2005); the subduction data, in blue is only pictured if MIS5e uplifted shoreline data are associated. The blue boxes display the three places where records are suitable to observe the uplift rate distribution along a trench perpendicular transect. B to D, zoom on areas of

special interest (see transects, sections 2.2 and 3.1); colours represent the uplift rate and the symbol (square/circle) represent the location of data less/more than 2degrees away from the trench.

2.2 Geological context of the South America, Cascadia and Japan-Korea margins

We found 3 places in the world where a repartition of terrace data from the fore-arc area toward the inner upper plate is available (Figure 2): the Juan de Fuca plate subduction under the Cascades, the Nazca plate subduction under South America and the Philippine Sea plate subduction under Japan and Korea. The Juan de Fuca plate subduction zone extends ~1000 km from the north of the Mendocino triple point to the south of the Queen Charlotte Island (51°N/130°W; 40°N/110°W). The subduction obliquity varies from 30° to the south to 15° to the north with a plate convergence rate estimated at 3.1 cm/y (Heuret and Lallemand, 2005). The age of the oceanic crust entering the trench is 10 My (Müller et al. (1997)). The tectonic setting of the overriding plate varies from slightly extensive to the south to mostly wrench faulting to the north (E1 to 0 as described thereafter; Heuret and Lallemand, 2005). The distance between the location of uplift data and the trench varies between 60 and 135 km (Figure 3B). The upper plate lithosphere elastic thickness is 10-30 km (Lowry et al. (2000)).

Coastal sequences on the Japan/Korea coasts are located above the Nankai-Ryukyu subduction zone where the Philippine Sea plate subducts below the Eurasian plate. This area extends latitudinally from Tokyo to Pyongyang, and from Honshu Island to the Tanega-Shima Island from E to W (40°N/126°E; 30°N/139°E). The obliquity of the subduction ranges from 20° to the west to 40° to the east (Heuret and Lallemand, 2005). Plate convergence in this area is 4.8 cm/y and the oceanic crust entering the trench is 35 My-old (Müller et al. (1997)). The tectonic setting of the Eurasian plate is slightly compressive on the whole studied segment (C1 as described thereafter) (Heuret and Lallemand, 2005). In this area, the distance between uplift data and the trench varies between 60 and 900 km (Figure 3A).

The South American area analyzed in this study is located along the ~6000 km-long Nazca subduction zone (from 5°N to 45°S, figure 2D). The Nazca subduction presents several particular zones of interest: the Carnegie ridge (0.5°N to 2°S; Gutscher et al. (1999a)), the Peruvian flat subduction (2 to 15°S, Gutscher et al. (1999b)), the Nazca ridge entrance into subduction (13 to 16°S; e.g., Machare and Ortlieb (1992), Hampel (2002), Espurt et al. (2008), Saillard et al. (2011)) and the Andean flat slab region which is located between 27 and 32° S (Yañez et al. (2001)). Plate convergence in this area is about 7 cm/y and the oceanic crust entering into the trench is 0 to 40 My-old (Müller et al. (1997); Heuret and Lallemand, 2005). Subduction obliquity is everywhere smaller than 20°. The South American plate tectonic setting presents a North to South compression gradient (Heuret and Lallemand, 2005), from compressive in Central Andes to neutral in Southern Andes; its elastic thickness possibly strongly varies from less than 10 km along the arc to >50 km in the foreland (Tassara (2005); Tassara et al. (2007)).

3. RESULTS

3.1 Distance to the trench

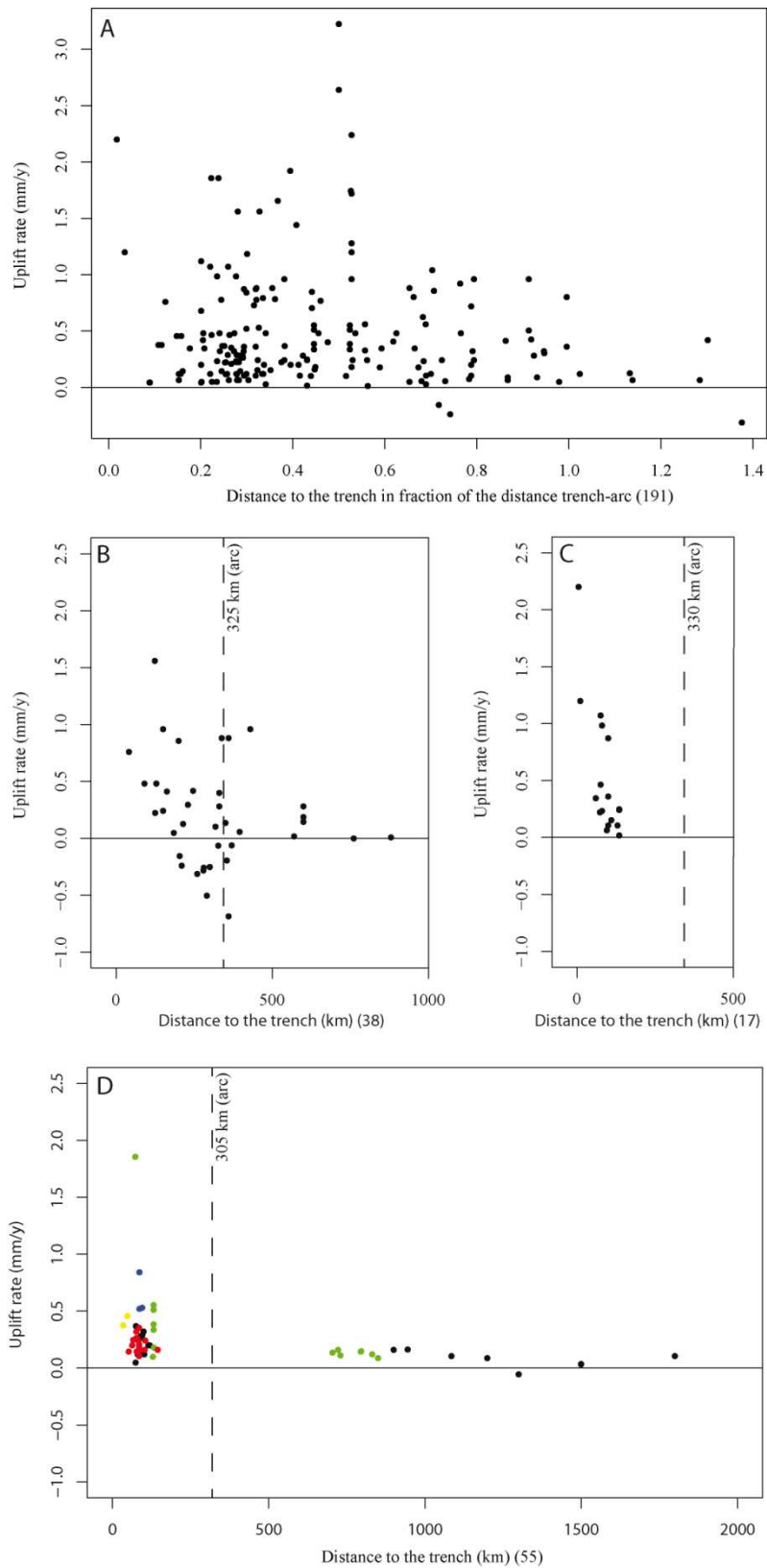


Figure 3. Uplift rate versus distance to the trench. *A: All data, expressed in function of the trench-arc distance. B: Philippine Sea plate subduction. C: Cascadia subduction. D: Nazca subduction; data located above the Peruvian and the Juan Fernandez flat subduction areas are shown in red (2-15°S and 27-32°S), over the Nazca Ridge in blue (13-16°S) and over the Carnegie Ridge in yellow (0.5°N-2°S). In addition green points are data from Patagonia (south of 32°S).*

The distance to the trench anti-correlates with uplift rates (Figure 3). Higher uplift rates are found at the closest points to the trench in every transect (Figure 3). Uplift rates generally decrease with the distance to the trench, with sometimes an area of slower uplift (e.g., negative values at ~300 km from the trench in Japan, Figure 3B). A direct correlation between the distance to the trench and the uplift rate is clear close to the trench, less clear farther. The overall relation is marked by a negative correlation coefficient of -0.12 (<-0.24 for each individual zone, table 3), quite significant (p-value of 0.03), which is meaningless considering the trend is non-linear (it becomes non-linear when considering points farther than the arc, Figure 3). We observe that the area of strong uplift is restricted to the first ~300 km from the trench, corresponding to the forearc area.

3.2 Slab dip

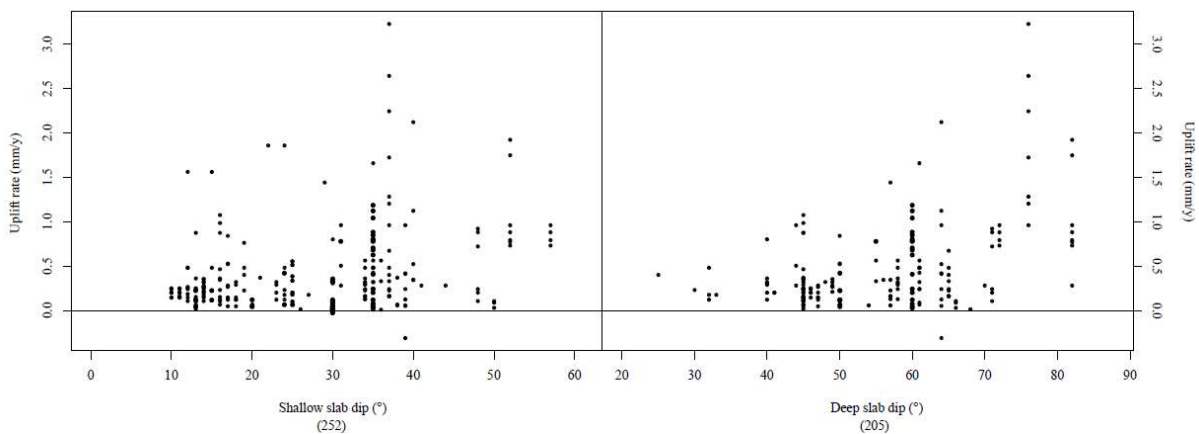


Figure 4. Left: uplift rate vs. shallow slab dip; right: uplift rate vs. deep slab dip, for the entire dataset.

Slab dip is measured at shallow depth (average dip between 0 and 125 km depth); and at greater depth, (average dip between 125 and 670 km depth)(Heuret and Lallemand, 2005). For both shallow and deep slab dip datasets, corresponding uplift rate appears to increase with slab dip (Figure 4). For the shallow slab dip the correlation is not obvious but marked by a slightly positive correlation coefficient of 0.32. The deep slab dip does not seem to influence the uplift rate below a critical value of ~60 degrees; above this value, there is a positive correlation between deep slab dip and surface uplift (Figure 4).

3.3 Tectonic setting of the overriding plate

Figure 5 shows the variations of uplift rate in function of the tectonic regime of the overriding plate. Heuret (2005) and Heuret and Lallemand (2005) describe the tectonic regime of the overriding plate using recorded focal mechanisms and classify it from the very compressive (C3) to very extensive (E3), 0 being neutral (no focal mechanism or wrench faulting). For this kind of qualitative parameter, it is not possible to calculate a correlation coefficient. We rather study them qualitatively, using boxplots (sometimes called box-and-whisker plots Tukey, 1977).

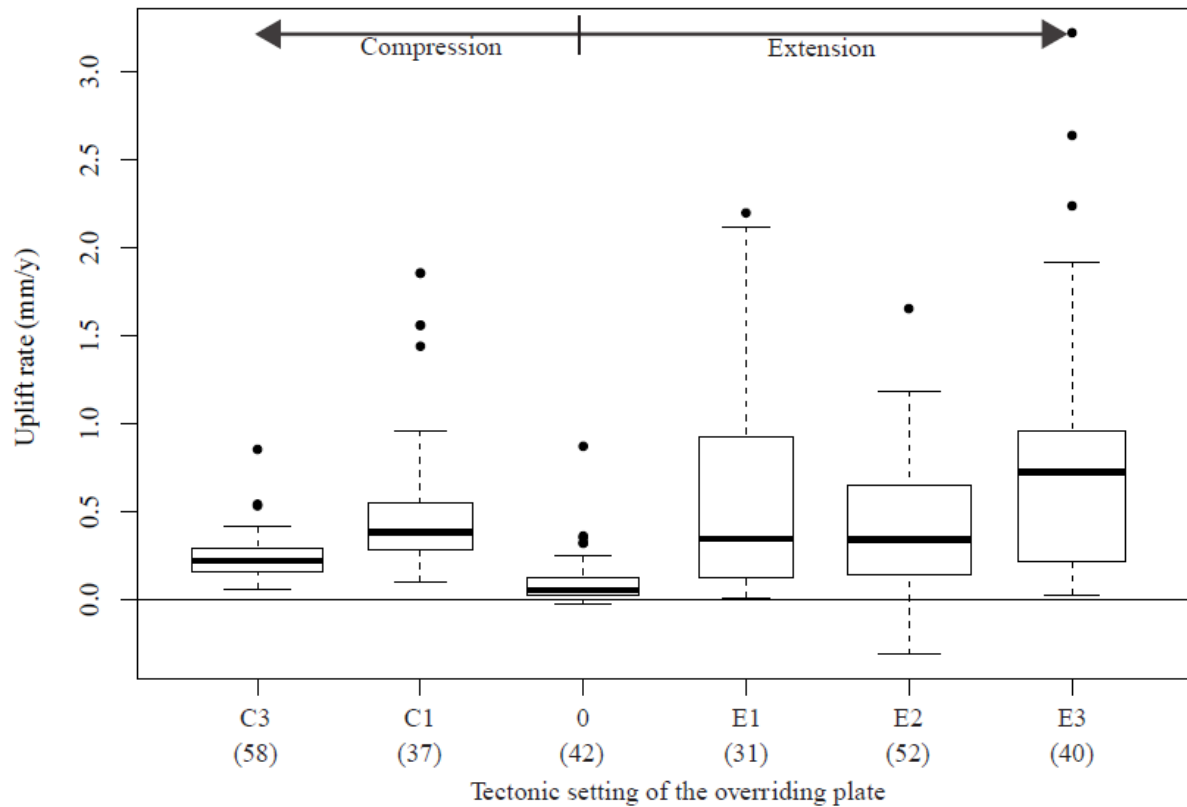


Figure 5. Uplift rate as a function of the overriding plate tectonic setting as defined by Heuret and Lallemand (2005). C3 is considered to be the most compressive state while E3 is the most extensional. C2 is absent because no C2 area is related to MIS5e marine terraces. Data number appears in parentheses under each box. Additional information can be found in Table 1. Horizontal bars indicate, from the bottom to the top, the first decile, the first quartile, the median, the third quartile and the ninth decile; dots are outliers. The horizontal size of the box is function of the data number.

Figure 5 shows a slight tendency to greater uplift rates when the tectonic setting is extensional rather than compressive. Noteworthy, intermediate settings (C1 to E2) are quite similar while the main trend comes from the extremes (C3 and E3 tectonic settings).

Table 1. Mean uplift rates (mm/yr) as a function of the tectonic setting of the overriding plate (C3 to E3) and for the position along the subduction zone expressed in “far” or “at” slab edge.

	C3	C2	C1	0	E1	E2	E3	At slab edge	Far from slab edge	Accretionary margins	Erosive margins
1 st quartile	0.144	NA	0.280	0.020	0.120	0.144	0.212	0.100	0.144	0.160	0.144
Median	0.208	NA	0.384	0.054	0.344	0.340	0.724	0.328	0.240	0.364	0.244
3 rd quartile	0.280	NA	0.550	0.120	0.928	0.652	0.950	0.872	0.364	0.780	0.406
Number of	58	NA	37	42	31	52	40	113	147	112	148

data											
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Figure 6 shows the relation between the uplift rate and the tectonic regime of the subduction margin. Accretionary margins show a higher magnitude (median: 0.36 mm/yr instead of 0.24 mm/yr for erosive ones) and a greater variability in uplift rate than erosive margins. The difference between those two regimes is somewhat significant: the median value for the accretionary margin (0.36 mm/y) corresponding to the third quartile value of the erosive margins (0.41 mm/y). The frequency of observation of uplift/subsidence in accretionary and erosive margins does not differ significantly (Table 2). This, together with the figure 6 pleads for a more variable vertical motion at accretionary margins than at erosive ones but without preferential (uplift/subsidence) direction.

Table 2. Number of data showing uplift, subsidence as function of the subduction margin tectonic setting. It reads as following example: the dataset of accretionary margins over 139 data, 26 (19%) record uplift and 2 (1%) record subsidence. There is no vertical motion data for the remaining 111 (80%) data.

	Accretionary	%	Erosive	%
Number of data	139		121	
Uplift	26	19	40	33
Subsidence	2	1	1	1
No associated data	111	80	80	66

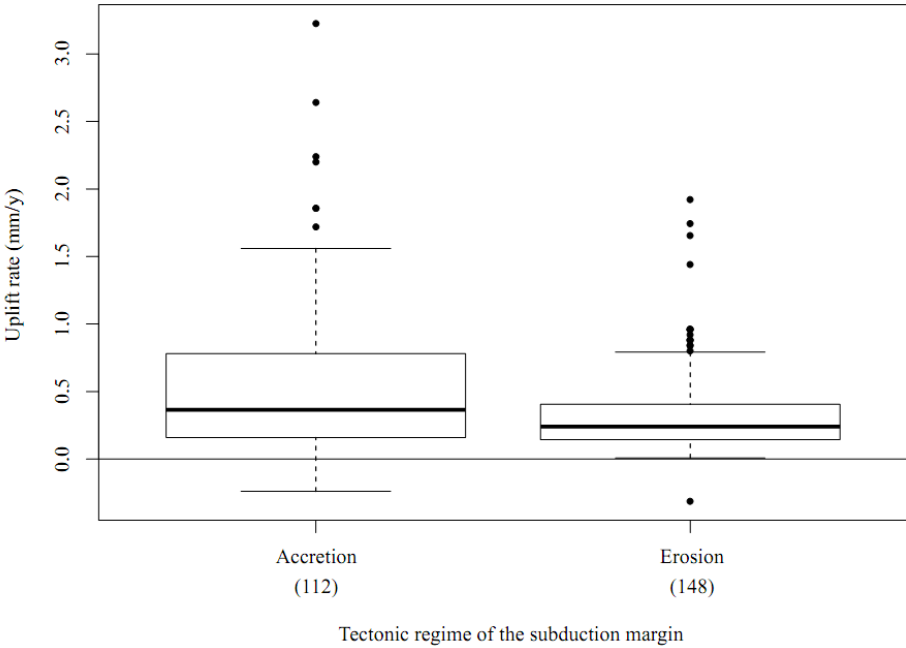


Figure 6. Uplift rate as a function of the subduction margin tectonic setting. See caption of Figure 5 for description of figure. Data number appears in parentheses under each box.

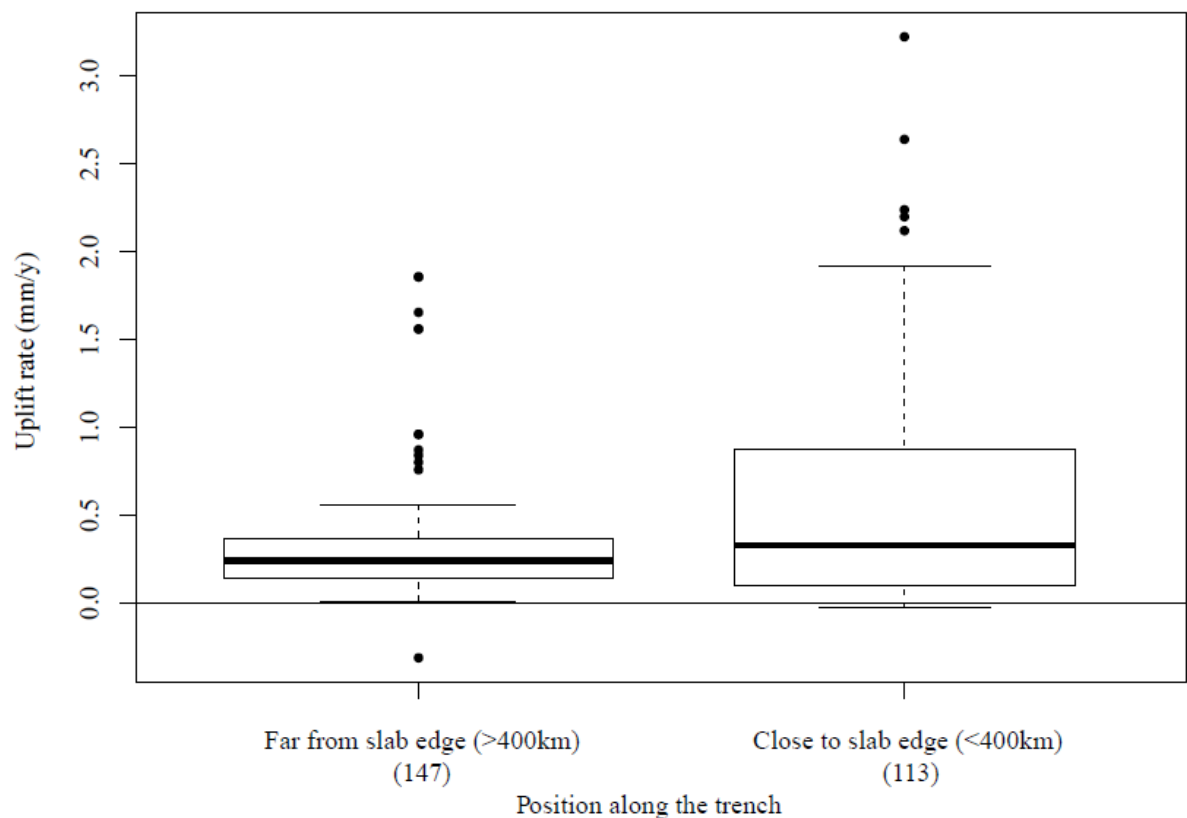


Figure 7. Uplift rate versus distance to a subduction edge. See caption of Figure 5 for description of figure. Data number appears in parentheses under each box. Additional information can be found in Table 1.

The position along the trench may be an important parameter as emphasized by the particular tectonic setting of slab lateral boundaries (Funiciello et al. (2004); Lallemand et al. (2005); Schellart et al. (2011)). According to Lallemand et al. (2005), a datum is considered to be at a subduction edge if it is closer than four degrees from the edge of the subduction. The uplift rate is shown as a function of the proximity of the measurement to a slab edge in figure 7. This figure shows that slab edges are characterized by higher uplift rates than elsewhere along the subduction zone. The difference is somewhat significant: the median value at slab edge corresponds to the third quartile value far from the edges (average uplift rate of 0.33 and 0.36 mm/y, respectively, Table 1).

3.5 Dynamic topography

Temporal changes in dynamic topography can offset sea level relative to the continent. A local decrease in dynamic topography leads to sea transgression while and conversely, an increase will produce continent emersion, regression, marine terrace fossilization and uplift. Thus we calculated the rate of change in dynamic topography through the last million years. Two models of dynamic topography change have been tested (Conrad and Husson (2009); Müller et al. (2008)). Combination of the 2 datasets (coastal uplift and dynamic topography) was performed in a 0.5 degree-bin because of the high apparent resolution of the dynamic topography models. Table 3 shows the comparison between the average observed uplift and the average vertical ground motion possibly due to changes in the dynamic support of the topography. Figure 8 shows the relationship between the rate of change in the modelled dynamic topography and measured uplift.

Table 3. Comparison of the observed uplift (Pedoja et al. (2011)) with modelled dynamic topography rates of change at the same locations (Conrad and Husson (2009); Müller et al. (2008)). The values are the average of 0.5 degrees bins regarding subduction zones. Note that the uplift observed from Pedoja et al. (2011) consequently differs from the average values of the entire set of observations (0.2227 mm/yr).

	Average uplift rate since the MIS 5e (mm/y)	Average uplift since the MIS 5e (m)
Observed uplift from Pedoja et al. (2011)	0.3533	44.16
Dynamic topography after Conrad and Husson (2009)	0.005	0.625
Dynamic topography after Müller et al. (2008)	-0.029	-3.625

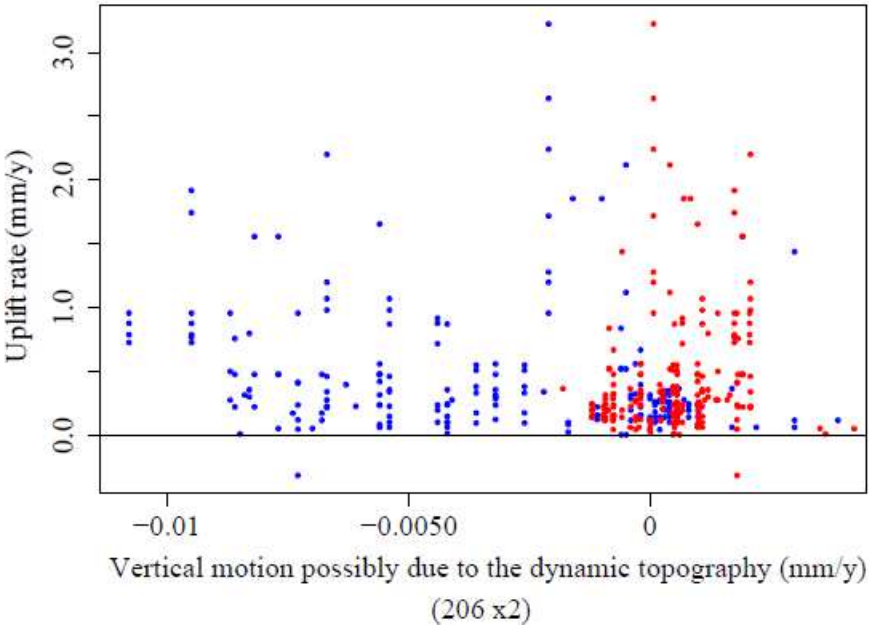


Figure 8. Uplift rate versus rate of change in dynamic topography for the last Ma.; in red, after Conrad and Husson (2009); in blue, after Müller et al. (2008).

According to the tested dynamic topography models, the average contribution to the vertical motion would be one order of magnitude underneath the observed uplift. In addition to this, figure 8 and the related correlation coefficient prevent from stating that there is a significant link between the dynamic topography and the observed coastal uplift rate. Last, the fact that most coastlines are uplifting suggests that dynamic topography shall not explain this phenomenon; indeed, uplift and subsidence must overall cancel each other and average to zero.

3.6 Other parameters

The other tested parameters do not show any trend with coastal uplift. Their correlation coefficient is too low to conclude on the robustness of any link between the measured uplift and tested geodynamic parameters. Among these, we explored the correlation of uplift rates with the commonly invoked overriding and subducting plate velocities, trench motion, convergence acceleration, subduction obliquity, oceanic crust age, interplate force and friction

force. Figure 9 summarizes the relationship between all the aforementioned parameters and the coastal uplift.

Plate and trench motion

The absolute plate motion comes from Steinberger et al. (2004). In Figure 9 A and B, positive (negative) values of subducting or overriding absolute plate motion indicate a movement toward (away from) the trench. The normal overriding plate velocity does not appear to be correlated with the uplift rate. Although the normal subducting plate velocity visually seems slightly correlated to the uplift rate the correlation coefficient is nearly null (~ 0.02) (Figure 9B). Finally, the convergence velocity is also surprisingly not correlated with terrace uplift (Figure 9A), even if we only take into account the trench-normal convergence velocity (Figure 9B). No correlation between trench motion and uplift rate is detectable (Figure 9C). In addition we evaluated the current change in convergence velocity for the plates involved in the Tonga-Kermadec, Japan, South America, Aleutian and Cascadia subduction (Figure 9D) by comparing a compilation of GPS derived, instantaneous velocity data (Sella et al. (2002)) to the integrated velocity for the last 3.2 Ma (NUVEL-1A, De Mets et al. (1994)). It is not possible to derive any correlation since the plate velocity change is distributed in patches. Anyway, the visual inspection does not plead for any correlation.

Obliquity

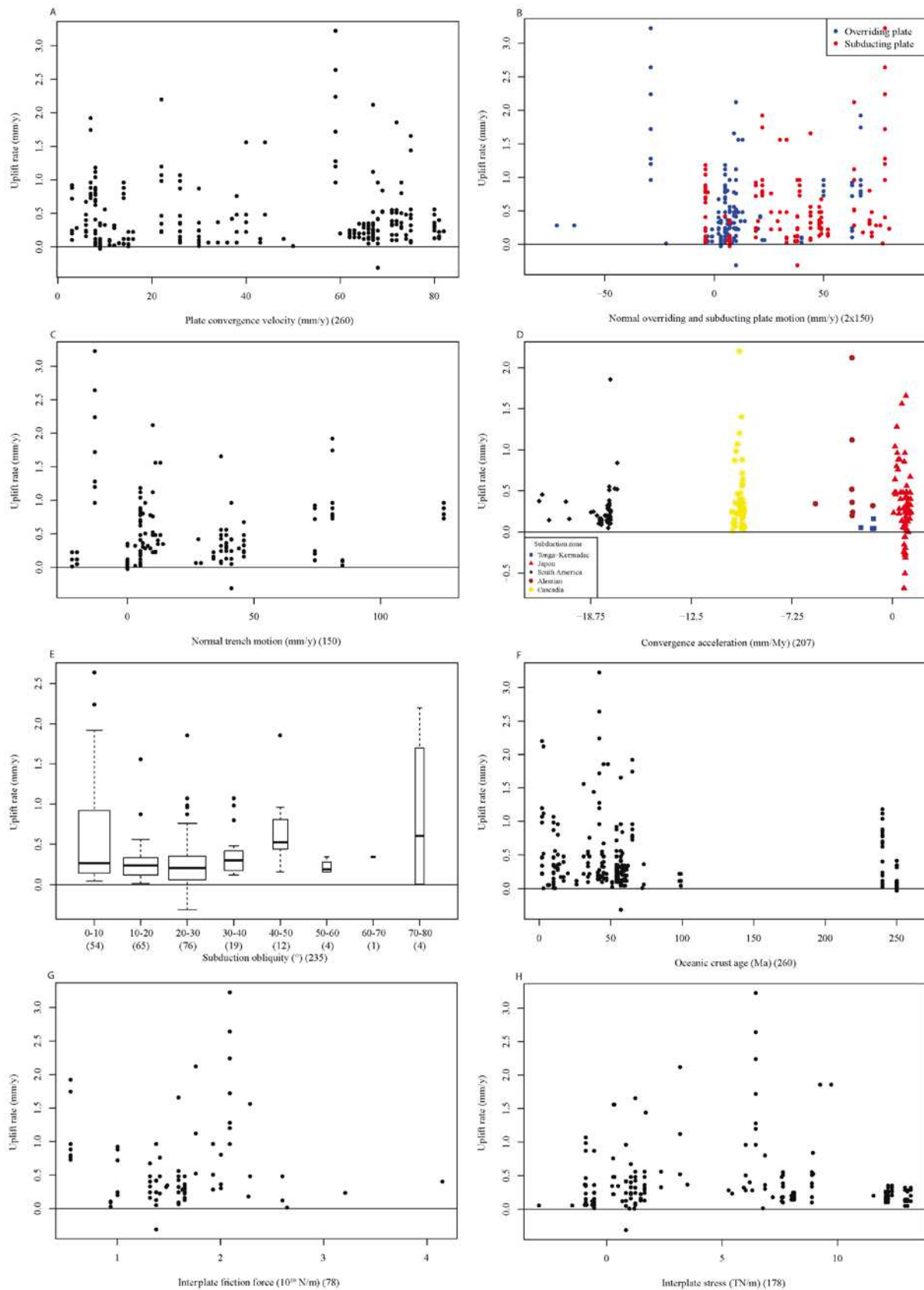
The subduction obliquity is calculated using DeMets et al. (1994) NUVEL-1A plate velocity model and Heuret (2005)'s trench azimuth (Figure 9E). An obliquity value of 0° indicates a perfectly orthogonal subduction. The dataset covers a wide range of obliquity values that are well distributed. No trend is observed except a possible slight increase towards normal convergence.

Oceanic crust age

The oceanic crust age is compiled by Heuret and Lallemand (2005) after Müller et al. (1997) and indicates the age of the oceanic crust at the trench. This distribution is not correlated with the uplift rate.

Interplate friction force and interplate force

Figure 9G shows the interplate friction force (*sensu* Lallemand, 1999) vs uplift rate. The interplate force is estimated by Husson (2012) by calculating the mean of the integrated mantle drag from both the overriding and subducting plate (M_d). There is detectable uplift rate correlation neither with the interplate friction force nor with M_d (Figure 9G & H).



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Figure 9. A: Uplift rate as a function of the convergence rate; B: Uplift rate as a function of the overriding (blue) and subducting plate (red) absolute motion; C: Uplift rate as a function of the absolute trench motion; D: Uplift rate as a function of the convergence variation for the last 3.2 Myrs, as the difference between NUVEL-1A and REVEL models, the different subduction zones are

pictured; E: Uplift rate as a function of the convergence obliquity; F: Uplift rate as a function of the oceanic crust age at the trench; G: Uplift rate as a function of the interplate friction force. H: Uplift rate as a function of the interplate force.

4. SUMMARY AND DISCUSSION

Table 4. Linear correlation coefficient between the tested geodynamic parameters and the coastal uplift. Bold lines indicate significant correlation (or anticorrelation) (p-values < 0.01).

Geodynamic Parameter	Correlation coefficient	Confidence interval (1 σ)	p-values
Subduction obliquity	0.019	[-0.04 0.082]	0.757
Oceanic crust age	-0.148	[-0.208 -0.087]	0.016
Normal Trench motion	0.072	[-0.0101 0.153]	0.382
Interplate force	-0.06	[-0.134 0.015]	0.427
Interplate friction force	0.03	[-0.085 0.144]	0.795
Overriding plate velocity	0.026	[-0.056 0.108]	0.748
Subducting plate velocity	0.239	[0.160 0.315]	0.003
Shallow slab dip	0.286	[0.219 0.350]	4.80x10⁻⁵
Deep slab dip	0.431	[0.368 0.490]	1.70x10⁻⁹
Convergence velocity	0.015	[-0.047 0.077]	0.805
Normal convergence velocity	0.203	[0.124 0.280]	0.012
Convergence acceleration since 3.2 Ma	0.051	[-0.018 0.120]	0.467
Distance to the trench (boxes)	-0.28	[-0.366 -0.189]	0.003
Distance to the trench, in fraction of the distance trench-arc	-0.119	[-0.189 -0.047]	0.031
Distance to the trench (South America transect)	-0.355	[-0.469 -0.229]	0.007
Distance to the trench (Japan-Korea transect)	-0.248	[-0.400 -0.088]	0.128
Distance to the trench (Cascadia transect)	-0.769	[-0.858 -0.637]	3.00x10⁻⁴
Dynamic topography (Conrad and Husson (2009))	0.191	[0.123 0.257]	0.005
Dynamic topography (Müller et al. (2008))	-0.229	[-0.294 -0.162]	9.10x10⁻⁴
Overriding plate tectonic setting	Gently correlated		
Position along the trench	Significantly correlated		
Tectonic regime of the subduction margin	Gently correlated		

Table 4 summarises the correlation between observed uplift rates and all tested parameters. Overall, most of the investigated geodynamic parameters do not seem to influence the uplift rate. We additionally observed higher uplift rates near lateral slab boundaries. Lallemand et al. (2005) observe that slab dip is generally higher at slab edge. Slab edges are also known to be places of high upper mantle toroidal motion, and eventually mantle vertical flow (e.g., Funicello et al. (2006); Guillaume et al. (2010); Schellart et al. (2007)). This leads to important variations in geodynamics at slab edges (Schellart et al. (2011)). Out of all the tested parameters, two parameters are highly correlated to uplift. First, slab dip shows some correlation with the uplift rate. Lallemand et al. (2005), in turn, show that the slab dip

correlates to the overriding plate tectonic setting. The higher the slab dip is, the more extensive the regime of the overriding plate will be. It is fully coherent with the fact the tectonic setting appears to be linked with uplift rate. But its effect is quite counterintuitive since uplift is possibly related to extension and low uplift to compression.

Due to the quasi-absence of subsiding data, it is not possible to tell if accretionary margins are more prone to uplift (and less prone to subsidence) than erosive ones, as shown by the works on subduction tectonic erosion (e.g., Lallemand et al., 1992). However, noticeable is the fact that erosive margins are not devoid of uplift zones despite the fact that intuitively, the forearc material loss associated with tectonic erosion is likely to promote subsidence. The normal faults that have been described in most erosive margins (e.g., Clift and Vannuchi, 2004) should be responsible of the observed uplifts. The difference between erosive/accretionary margins is best expressed in terms of variability of the uplift magnitude, greater for accretionary margins. Large uplifts at erosive margins are possibly inhibited by the forearc material related to the erosive processes.

Second and maybe the best correlated parameter to uplift is the distance to the trench which shows in 3 particular transects a rapidly slowing uplift with the distance: we evaluated the characteristic distance of action of the trench to be ~ 300 km, i.e., the trench-arc distance.

The main limitation of our approach is the small number of sites evidencing coastal subsidence, partly because measuring subsidence is more difficult than measuring uplift, partly because it is less common. For example, this excludes the Sunda subduction zone due to a lack of observational data. Even if we consider the record is correct for uplifting areas, the input of subsiding areas could enlarge the range of parameters to observe. For example, extrapolating the regression observed for steeply dipping slabs toward gentle dips predict negative values, i.e., subsidence. This will have to be investigated.

Another limitation is that “simple” subduction zones associated with a coastline prone to record fossil shores are not so frequent. For example the Tonga-Kermadec and Marianas subduction zones are ocean-ocean subduction zones, with limited subaerial exposure in which vertical displacements can be measured.

A possible reason for the poor correlation between uplift and the geodynamic parameters of subduction zones is that subduction zones correspond to a long-term phenomenon that lasts tens of Myrs. In contrast, the analysis of terraces associated with the MIS5e isotopic stage evidences short term ($\sim 10^5$ years) vertical displacements. Subductions zones evolve slowly through time and their morphology should be in equilibrium with the main geodynamic parameters such as the convergence velocity or the friction at the plate interface. One can expect that short-term coastal uplift (or subsidence) is mostly due to changes in these parameters. We thus explored the variation in subduction velocity and the variation in the dynamic topography but here again we fail at finding any correlation. Finally, the absence of correlation between geodynamics and vertical motion may plead for a local effect (or short in time) causing strong vertical motion. These local effects are preferentially due to roughness of the subducting plate as we illustrate by the examples in the following.

(1) In fact, the high uplift rate of E3 (strongly extensive) tectonic setting can come from the position of E3 data close to slab edges (38 data over 40). The north New Hebrides (Vanuatu) subduction zone is typical from E3 tectonic setting. It is marked by back-arc opening but also by many plateau/ridges entering into subduction (Taylor et al. (2005)), probably causing a high variability in uplift rates. They range 0.004-1.896 mm/y, with a median uplift rate of ~ 0.756 mm/y, close to the median value for E3 (Figure 5 and Table 1). This subduction zone clearly questions the origin of the high uplift rate: the extensional setting, the proximity to slab edges, the subduction of asperities like the d’Entrecasteaux ridge (Collot et al. 1985; Taylor et al. 2005), or a slab break-off (Châtelain et al. 1992)? A distribution of decreasing

uplift rates westward from the trench to the back arc basins is also clearly documented here (e.g., Collot et al. 1985; Châtelain et al. 1992; Taylor et al., 1987; 2005).

(2) The Nazca plate subduction under South America is characterized by a C3 tectonic setting and a moderate uplift rate (median at 0.176 mm/y and range 0.024-0.816 mm/yr; see also the transect Figure 3); none of the points is at (or near) a slab edge. A particular characteristic of this subduction zone is the entrance of some aseismic ridges into the subduction zone (Gutscher et al., 2000), causing the highest uplift values of this area (e.g., Machare and Ortlieb (1992); Martinod et al. (2013); Regard et al. (2010); Saillard et al. (2011)). Of particular interest is the Nazca Ridge entering into the subduction zone in central-south Peru (Machare and Ortlieb (1992); Regard et al. (2010); Saillard et al. (2011)). The contact point of the ridge with the trench is moving southeastward (Hampel (2002), Espurt et al. (2008)). North of this contact point the coastline is subsiding (cf. in Lima, Leroux et al. (2000)), after the Nazca Ridge passed through. Thus the slab buoyancy appears to be a major driving parameter for vertical motion in Peru. Another important fact is that if this subduction zone is characterized by Heuret (2005) as strongly compressive (C3), we note that the forearc behaves differently: the margin is believed to be tectonically erosional (e.g., Lallemand et al. (1992)) and it is sometimes extensional, in particular in the entire northern part of Chile (e.g., Gonzalez et al., 2003).

Finally, these examples show that the uplift rate recorded by ancient shorelines mostly concerns the forearc, whose tectonic setting (~100-300 km from the trench) is sometimes disconnected from that of the back-arc area as recorded by Heuret and Lallemand (2005) (~300-500 km). We can hypothesize that the upper plate quickly responds vertically to external factors, as every change in geodynamical setting. Consequently, the subduction zones are generally close to equilibrium with the external forces causing stability (or slowly evolving) in terms of vertical motion. Superimposed on to this geodynamic setting a ridge entrance into subduction is prone to cause significant uplift, maybe more efficiently if the far field upper plate tectonic setting is extensional. That is why high uplift rates are transient and cannot last more than a couple of hundred of thousand years: figure 10 shows that high uplift rates are only found in places uplifting for a short period of time (<1-2 My). The ridge effect is important when the ridge slides along the trench as the Nazca Ridge does (or when a new ridge enters into trench). On the contrary a ridge entering the subduction at the same point for a long time would not cause a temporal change: a good example is the Juan Fernandez Ridge in Chile.

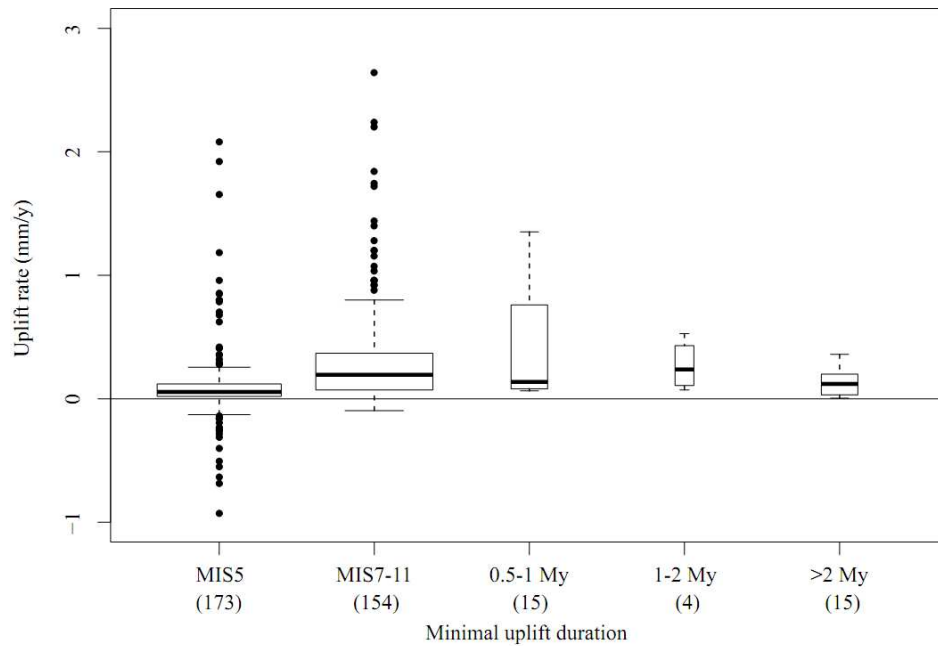


Figure 10. Uplift rate since MIS5 as a function of the observed duration of uplift (maximum age of uplifted shorelines).

5. CONCLUSION

We explored the relationship between the observed coastal late Pleistocene uplift rates and various geodynamic parameters. This statistical study shows that most of the geodynamic parameters are not related to coastal uplift magnitude. The distance to the trench presents a slightly correlated signal with the coastal uplift. It is possible this effect ends at ~300 km from the trench, which approximately corresponds to the position of the volcanic arc. Other slight correlations have been found between uplift and slab dip, position along the trench and overriding plate tectonic regime.

The main message is that subduction zones are characterized by rapid vertical motion (uplift but maybe also subsidence as stated in introduction), only part of which being satisfactorily explained by the geodynamic setting. Over this global setting, strong vertical motion is due to smaller scale heterogeneities (one can call roughness) of the subducting plate, one typical example of roughness being a subducting aseismic ridge. Asperity affects more efficiently the areas close to the trench.

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